HELLENIC SHELF: LATE QUATERNARY TECTONICS, SEA-LEVEL CHANGES, SEDIMENTATION AND GEO-HAZARDS.

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Number of words: 8,467

Abbreviated title: The Hellenic Shelf

Abstract

The Hellenic shelf lies within and around the Aegean micro-plate, which is one of the world’s most seismically active areas and has experienced extreme tectonism throughout Quaternary time. This activity together with eustatic sea-level changes and water circulation patterns over the same time period, control the overall configuration of the Hellenic shelf, the rates of uplift and subsidence and determines, the sediment supply and depot centre as well as the sediment transfer processes. The above mentioned geological processes are the causative factors for the frequent occurrence of a variety of geological hazards such as active faults, submarine gravitational mass movements, tsunami and active gas seeping from the seabed.

The Hellenic shelf lies within and around the Aegean micro-plate, which is one of the most rapidly deforming areas in the world (Fig. 1) (Jackson, 1944). It comprises the Aegean, an epicontinental sea (Stanley & Perissoratis 1977) and the western margin of the Hellenic peninsula in the Ionian sea (Fig. 1). The Hellenic shelf can therefore be considered as a natural laboratory for the study of continental tectonics such as fault geometry and kinematics, basin formation and filling processes, sediment transport processes and quantitative modeling of erosion and deposition in relation to the tectonic uplift and subsidence.
Furthermore, the Hellenic shelf, due to the strong and frequently occurring earthquakes, can be considered as an ideal area for the study of a variety of geo-hazards such as active faults, submarine landslides, tsunami and active gas seeping from the seabed and, their causative mechanisms.

**Tectonic framework**

The Quaternary active deformation of the Hellenic Peninsula and the surrounding Ionian and Aegean seas shelf is the result of (McClusky et al. 2000; Taymaz et al. 1991, 2007): (i) the westward movement of the Anatolian micro-plate between the North Anatolian Fault (NAF) and the East Anatolian Fault (EAF) (Fig. 1a), at a rate of 20 mm/year, in response to the northward collision of the Arabian plate into the Eurasian plate, (ii) the southwestward movement of the Aegean plate at 30 to 32 mm/year and its collision with the Apulian micro-plate to the west and the African plate to the south and southwest (Fig. 1a). The effect of these plates movement during Quaternary is reflected in the spatial and temporal variation of the observed fault pattern and stress field all over the above mentioned area (Doutsos and Kokkalas 2001, Kokkalas et al. 2006).

**The Hellenic shelf**

The Hellenic shelf along the western margin of the Hellenic peninsula in the Ionian sea lies over the Hellenic Compressional Belt in the outer arc domain (Fig. 1). The Hellenic shelf in the Aegean, an epicontinental sea (Stanley & Perrisoratis 1977), overlies the continental crust of the Alpine Hellenides tectonic fabric of folds, thrust faults and sutures formed during the Alpine orogeny. The complex deformation of the crust in the Hellenic shelf, which lies within and around the Aegean micro-plate, during Quaternary may be the result of: (1) slab retreat along the subduction zone and consequent back-arc extension, (2) collapse of an overthickened crust, (3) the westward escape of Anatolia along its plate boundaries and (4) differential rates of convergence between the NE directed subduction of the African plate in relation to the Anatolian plate (Taymaz et al. 2007).
The Hellenic shelf in the Ionian Sea

The Hellenic shelf in the Ionian sea, based on the major morphotectonic features that shape the Hellenic Compressional Belt, can be divided into three segments: (i) the Illyrian Folding belt (IFb), (ii) the Lefkada–Kefallinia escarpment (LKe) and (iii) the Hellenic Subduction zone (HSz) (Fig. 1).

The Hellenic shelf in the Ionian sea extends up to the 120m isobath, has a width ranging from 3 to 50 km and dips seawards at an average gradient of 0.6°.

The shelf along the Illyrian Folding belt (IFb) runs along the coastal zone of southern Albania and continues south along the margin of western Greece as far as the northern end of the Lefkada Island (Fig. 1). It overlies the Apulian, pre-Apulian and Ionian zones of the external Hellenides (Monopolis & Bruneton 1982, figs 1 and 3), which is characterized by NE-SW trending active shortening with folding, reverse faulting and thrusting (Monopolis & Bruneton 1982; Hatzfeld et al. 1995).

The flatness of the seafloor and the absence of any compressional features that affect the sedimentary cover, suggest that the underlain reverse faults and thrusts do not affect the seafloor (Ferentinos & Papatheodorou 1994). The slope is narrow with a gradient of between 1.6° and 5.7° and is covered by a thin veneer of unconsolidated sediments (<20 m) and is probably the morphological expression of the thrust front (Monopolis & Bruneton 1982, fig. 4a).

The Hellenic shelf along the Lefkada–Kefallinia escarpment segment, develops between the island (Lefkada, Kefallinia and Zakynthos) chain and the Greek mainland (Fig. 1). The shelf has a width ranging from 25 to 50 km. It overlies the Apulian, pre-Apulian and Ionian zones of the External Hellenides (Monopolis & Bruneton 1982).

The shelf morphology is dominated by two narrow, elongated, parallel to the external Hellenides, NW-SE to NNW-SSE trending depressions about 60 and 50 km long each; the Zakynthos depression reaching water depths of 520 m and the Kefallinia depression reaching water depths of 340 m, separated by a NNW-SSE trending ridge (Brooks & Ferentinos 1984, figs 6 and 7). This morphological configuration of the shelf reflects the complex deformation of the underlain Pre-Apulian and Ionian zones, which is controlled by reverse
faulting, thrusting and associated diapirism, due to the active NE-SW compression (Monopolis & Bruneton 1982, figs 3 and 4; Brooks & Ferentinos 1984, figs 6, 7 and 8).

The slope develops to the west of the islands and is associated with the Lefkada-Kefallinia escarpment which is about 2 km in height and is characterized by steep slopes ranging between 9.4° and 18.4°. It is the surface expression of the Lefkada–Kefallinia transform fault (Kokkinou et al. 2006, figs 6 and 7).

The Hellenic shelf along the Hellenic Subduction zone in the Ionian Sea, runs from Zakynthos Island to the south-western tip of the Peloponnesus and from there, continues to western Crete (Fig. 1). The shelf, is narrow (2.5 to 5 km), flat and extends up to the 120m isobath (Papanikolaou et al. 1988; Poulos et al. 2002). The shelf configuration is controlled by active faults trending NW- to NNW- and NNE-. The NW to NNW faults are responsible for the formation of NW-SE and NNW-SSE trending onshore and offshore elongated basins associated with the Messiniakos, Lakonikos and Argolis gulfs in southern Peloponnesus (Kokkalas et al. 2006, fig. 6) and the formation of two elongated offshore basins in western Crete, the Gramvousa and Kissamos basins (Fig.1c). A sparker profile, which runs across the two basins (Fig. 2) shows that the faults are closely spaced with steep fault planes that seem to converge with depth and displace individual tilting blocks. This fault pattern indicates oblique normal faulting. Taking into consideration the similarities they exhibit with the nearby faults on land (Kokkalas et al. 2006, fig. 6), it is concluded that these two offshore faults are oblique normal faults. The NNW- and NNE-trending faults borders the Maleas-Kythera-Crete ridge (Fig. 1c) and are responsible for its formation.

The study of 3.5KHz., high resolution, seismic profiles across the Ionian sea shelf, shows that the sediments deposited on the shelf overlie an erosion terrace cut by the rising sea, during the post Last Glacial transgression (18 ka BP) (Ferentinos & Papatheodorou 1994; Poulos et al. 2002, fig. 3; Ferentinos et al. 2012, fig. 6). These deposits correspond to the late Holocene Highstand System Tract (HST) and are characterised by a typical sigmoid progradational configuration consisting of: (i) discontinuous and discordant reflectors along the present-day river mouths, representing topsets, (ii) stratified inclined
reflectors, representing foresets and (iii) stratified horizontal reflectors, representing bottomsets (Poulos et al. 2002, fig. 3). The HST sedimentary sequence may have or may have not prograded across the shelf edge draping the shelf and slope.

The Hellenic Shelf in the Aegean Sea

The Hellenic shelf in the south Aegean sea occupies the inner arc domain of the Hellenic Arc, which consists of three morphotectonic units: the Cretan outer arc, the Cretan Sea fore-arc basin (CSb) and the Attico-Cyclades Volcanic Arc/Metamorphic Core Complex continental platform (ACp) (Fig. 1)

The south Aegean is dominated by a complex tectonic regime (Fig. 1) (Mascle & Martin 1990; Kokkalas & Doutsos 2001; Piper & Perissoratis 2003; Kokkalas et al. 2006; Kokkalas & Aydin 2012) characterised by: (i) an older set of mid to late Miocene E-W trending normal faults as a result of pull-apart between stable Europe and south moving Aegean micro-plate, (ii) a younger set of Pliocene NE-SW trending strike slip faults that corresponds to the time when the North Anatolia Fault propagated to the north-eastern Aegean and (iii) a set of mid Quaternary to present NNW-SSE to NNE-SSW trending normal faults in south-western Peloponnesus and south-western Aegean sea and ENE-WSW trending strike-slip faults in south-eastern Aegean sea, which created and/or reactivated secondary normal faults trending NNE-SSW, NNW-SSE and E-W (Mascle & Martin 1990, fig. 1; Piper & Perissoratis 2003, fig. 20; Kokkalas and Aydin, 2012, fig. 13). The above mentioned tectonic regime controls the formation of the Cretan Sea, which is an arc shaped basin consisting of a fault controlled chain of smaller basins trending NW-SE and WNW-ESE in the west, W-E in the central part of the arc and NE-SW in the east and ranging in depth from 1400 to 2658 m (Bartole et al. 1983, fig. 4) and of the Ikaria, Amorgos, Anafi and Kos basins in the eastern part of the Attico-Cyclades continental platform (ACp). The Cretan sea basin is bordered to the south by the Cretan shelf, which extends to the 120 m isobath and to the north by the Attico-Cyclades continental platform shelf, which extends to 120 m isobath. Both shelves are generally flat and are covered by a thin sequence (<~20 m) of layered unconsolidated sediments, which overlie the post Last Glacial transgression erosion surface (18 ka BP.) (Ferentinos &
Papatheodorou 1993; Ferentinos & Papatheodorou, 1995). This sequence corresponds to the late Holocene Highstand System Tract (HST). In the Attico-Cyclades shelf the presence of between two and four successive oblique progradational delta sequences at the western (Kapsimalis et al. 2009, figs 2, 3 and 4), southern (Piper & Perissoratis 2003, fig. 11) and eastern (Lykousis et al. 1995, fig. 2) margin, suggest that the shelf continuously subsides. These successive progradational delta sequences were deposited during low sea level stands during middle and late Quaternary glacial stages corresponding to the Oxygen Isotopic Stages (I.O.S.) 2.2 (18 ka BP), 6.2 (146 ka BP), 8.2 (250 ka BP) and 10 (340 ka BP). Similar stacked delta deposits have been found in Argolikos gulf (Piper & Perissoratis 2003, fig. 4) and in Izmir bay in eastern Aegean (Aksu et al. 1987, figs 6 and 7) suggesting that the shelf over the Cyclades subsides. The estimated subsidence rates range from 0.35 to 0.90 m/ka and vary in space and time; (Lykousis et al. 1995, 2007; Lykousis 2009, table 1).

The Hellenic shelf in the northern Aegean sea includes the Macedonian-Thracian shelf (M-Ts) in the north and the Limnos continental platform (S-Lp) in the south separated by the North Aegean Trough (NAT) and the Skyros-Edremit Trough (S-Et) (Fig. 1). The two troughs are considered morphological expressions of the westward extension of the northern and southern branches of the North Anatolian Transform Fault. The North Aegean Trough (NAT) is bounded to the north by the Macedonian-Thracian shelf (M-Ts) and to the south by the Sporades-Limnos continental platform shelf (S-Lp). It is about 346 km long and 30 km wide and is divided into two basins; the Saros basin (Sab) to the east, trending ENE-WSW and the Sporades basin (Spb) to the west, trending NE-SW (Brooks & Ferentinos 1980; Stanley & Perissoratis 1977). The Saros basin is bounded to the north and south by the Ganos right strike slip fault zone (Koukouvelas & Aydin 2002). The Sporades basin at the eastern end is bounded to the south by the Sporades/Limnos fault zone (S-Lfz) and to the north by the Sithonia/Athos fault zones (S-Afz), which are characterized by normal and right strike slip components (Rousos & Lyssimachou 1991, figs 2, 5, 6 and 7; Koukouvelas & Aydin 2002, figs 7 and 8). The faults are active as they offset Pleistocene and Holocene strata (Fig. 3). The Sporades basin at the western end changes direction from NE-SW to
WNW-ESE due to the activity of listric faulting (Fig. 1c). The basin to the
south-southwest is flanked by a series of active synthetic listric faults dipping
to north northeast (Ferentinos 1992, fig. 2; Laigle et al. 2000, fig. 2), whilst to
the north-northeast is flanked by an active listric fault which dips to the south-
southeast (Ferentinos 1992, fig.2). Between the NNE and SSW dipping listric
faults an anticlinal structure trending WWS-ESE is formed due to space
reduction caused by the opposite moving hanging wall of the listric faults. Air-
gun seismic profiles across the anticlinal structure show evidence of a close
association of folding and faulting (Brooks & Ferentinos 1980, fig. 7). These
faults offset the seafloor and face upslope, are closely spaced and form
individual tilted blocks. The wedged-shape layers in each block indicate
growth faulting (Fig. 4). The Macedonian-Thracian shelf has a width ranging
from 22 to 34 km. Since the late Quaternary the shelf has undergone a
tectonic subsidence. At the western end, the Thermaikos shelf is subsiding
south-southeastwards for the last 450 ka BP under the influence of the listric
faults flanking the basin to the south-southwest. The Thermaikos shelf
subsides at an average rate of 0.93 m/ka as it is indicated by the presence of
a sequence of deltaic units with a cumulative thickness of 600 m. The older
unit (I.O.S. 12) now underlies a water depth of 700 m whilst the younger
(I.O.S. 2.2) is under the water depth of only 120 m (Lykousis 1991 a, figs 4
and 8). At the eastern end, the shelf edge subsides at an average rate of
between 0.3 to 0.5 cm/ka as it is indicated by the displacement of a delta
progradation sequence which corresponds to the I.O.S. 6 (146 ka BP) low
stand to a deeper water depth due to faulting (Piper & Perissoratis 1991, fig.
9). The Sporades-Limnos continental platform consist of Oligocene to
Miocene volcanic rocks and Eo-Oligocene sedimentary rocks and is fault
bounded to the north and south by ENE-WSW trending faults (Mascle &
Martin 1990).

Late Quaternary sedimentation

Over the last five decades a large number of papers have been published
relating to the: (i) the character of surficial sediments (Lykousis et al. 1981;
Anagnostou et al. 1993; Anagnostou et al. 1998; Volakis & Anagnostou 1993;
Giresses et al. 2003; Poulos 2009) (ii) transport processes and deposition

**Surficial Sediments**

The unconsolidated surficial sediments covering the seafloor of the Hellenic shelf, in terms of texture and composition, are mainly terrigenous muds (M) and sandy muds (sM) (Poulos 2009, Figs 6, 8 and 9). In the shelf areas, where the riverine input is minimal, the surficial sediments are mainly muddy sands (mS) and sands (S) of biogenic origin (Anagnostou et al. 1993; Anagnostou et al. 1998; Volakis & Anagnostou 1993; Georgiadis et al. 2009). The terrigenous sediments found on the shelf are considered to be late Holocene prodelta sediment sequences draping the shelf (Lykousis & Chronis 1989; Piper & Perissoratis 1991; Poulos 2002; Lykousis et al. 2005). This sedimentation has been active since 6000 yrs BP when the sea attained its present sea-level (Lykousis & Chronis 1989; Piper & Perissoratis 1991; Lykousis et al. 2005). The increased amount of sand observed in the mid and outer shelf represents either lowstand deltaic sediments or reworked shallow coastal sands deposited during the early stages of the postglacial transgression 18-6 ka BP (Piper & Perissoratis 1991; Lykousis et al. 2005).

The biogenic sediments that cover the Attico-Cyclades and Sporades-Limnos continental platform consist mainly of coraline algal debris (Lithothannium and Phymatolithon sp.), coral debris (Derdrophylia sp.) and mollusc shell debris (Anagnostou et al. 1993; Anagnostou et al. 1998; Volakis & Anagnostou 1993; Georgiadis et al. 2009). In the Attico-Cyclades continental platform shelf, in water depths between 40 and 120 m, the floor is covered by patches of coralligenous formations that have growing since the beginning of the postglacial transgression (Georgiadis et al. 2009, fig. 2, 3, 4, 5 and 6).

Further offshore, over the tectonically controlled deep basins, which dissect the Hellenic shelf, hemipelagic sedimentation prevails in the last 6000 years. In the northern Aegean, in the settling matter, the terrigenous content is the dominant one whilst in the southern Aegean the biogenic content is more pronounced (Poulos 2009). This hemipelagic sedimentation, in almost every
basin, is often interrupted by gravity flows (Giresse et al. 2003; Rousakis et al. 2004; Geraga et al. 2000, 2005).

Sediment Transfer Processes
Sediment transfer processes from land to shelf, then to slope and subsequently to the deep basin seafloor which operate in the Hellenic shelf, can be synoptically demonstrated by the following three key examples, dealing with three different shelf environments. The first demonstrates the transfer processes, which operate in a wide and low gradient (less than 1°) shelf with high river sediment input (Thermaikos Gulf), the second demonstrates the transfer processes that operate in a very narrow low gradient shelf with episodic but high river discharge (Corinth Gulf) and the third an island complex shelf with restricted sediment supply (Attico-Cyclades continental platform.

(i) The Thermaikos Gulf shelf at the western end of the Macedonian-Thracian shelf is about 12 km wide and is flat with a gradient less than 1°. At the Gulf head three major rivers discharge an annual mean average of 25x10⁶ tons/year sediments. A typical prodelta wedge showing sigmoid progradational configuration has developed in the Thermaikos Gulf shelf during the last 6000 years when the sea attained its present level (Lykousis et al. 2005, figs 2 and 5). The coarser material is deposited at the mouth of the rivers as topsets, the finer material at the delta slope as foresets and the finest is blanketing the shelf and slope seafloor as bottomsets (Lykousis & Chronis, 1989, fig. 3; Lykousis et al. 2005). The thin sediment blanket that covers the shelf overlies coarse sediments deposited during the post last glacial transgression from 18 to 6ka. These deposits overlie river deposits accumulated between 22 and 18ka when the sea level was standing 120m below present during the Last Glacial Maximum. At the shelf-edge at a water depth of 200 m at the seafloor deltaic sediments outcrop corresponding to the I.O.S 2.2 that is when the sea level was standing at 125m below the present (Lykousis & Chronis 1989, fig.6). The approximate 30m thick layered sedimentary sequence that drapes the slope consist of an upper layered unit that correspond to the late Holocene Highstand System Tract, which conformably overlies a lower unit that corresponds to the IOS 2.2 Lowstand
(Lykousis et al. 2005, fig.9). The bulk of the sediment deposited over the outer Thermaikos Gulf shelf and slope over the last 6000 years, is composed of terrestrial sediments supplied by the rivers and the surrounding land masses. These sediments have been dispersed over the shelf and slope via nepheloid layers and deposited as hemipelagic muddy deposits, as can be deduced from the distribution of silts, clays and clay minerals (Lykousis & Chronis 1989; Lykousis et al. 2005). Nepheloids layers have been observed to form at present time at surface, and bottom waters depths all the way from the river mouths down to the shelf edge and then to the slope (Durrieu de Mardon et al. 1992, fig. 2, 5 and 9; Karageorgis & Anagnostou 2001, fig. 2). At the same time dense water has been observed to form at the head of the Gulf in winter, by cooling during extremely cold northerly winds (Estournel et al. 2005). This dense water sweeps the sea seafloor across the shelf and slope with speeds of between 5 to 15 and 30cm/s in the shelf and slope respectively, transporting sediments to the basin.

(ii) The southern margin of the Corinth Gulf is the surface expression of step-like active fault segments (Stefatos et al. 2002). The margin is characterised by a narrow (less than 2km) low gradient shelf and a steep slope with a gradient ranging from 10 to 40. The slope is segmented and related to fault escarpment. The study of high resolution seismic profiles shows that the shelf and slope are dissected by canyons, which are either directly connected to ephemeral rivers on land, or begin at the shelf edge. The former directly transports the river sediment load to the basin floor as density flows and turbidity currents whilst the latter delivers sediments from the shelf to the basin floor, through retrogressive slumping at the shelf-edge and subsequent disintegration of the slumped masses to debris/mud flows and turbidity currents (Ferentinos et al. 1988). These processes control the sediment transport from land to shelf and deep basin seafloor during low and high sea level stands resulting in an average sedimentation rate of between 100 and 190 cm/ka (Lykousis et al. 2007; Bell et al. 2009). The mass movements are triggered by earthquakes, heavy rain and stormy weather and occur at least once every two years. Individual events can occur in two or more localities simultaneously along the shelf and slope and the sediment masses can travel between 2 and 6 km on the basin floor (Ferentinos et al. 1988).
The Attica–Cyclades continental platform shelf constitutes a shallow water barrier, with an average depth of 100 m, which separates the northern Aegean from the southern Aegean. The outflow of the low salinity Black Sea Water mass on the surface, the underlying Levantine Intermediate Water mass and the occasional cold, dense water masses forming during the dry, cold winter periods, lead to the generation of an intensive current which flows over the continental platform shelf through the island straits to the southern Aegean Sea. The current crossing through the straits sweeps all the sediments from the seafloor, leaving only outcrops of barren rock (Lykousis 2001, figs 2, 7 and 10). At the Straits between the islands the current can reach speeds of up to 0.8m/sec (Ferentinos & Papatheodorou 1993) transporting sediment from the northern Aegean Sea upslope over the plateau and then downslope to the southern Aegean Sea (Ferentinos & Papatheodorou 1993, fig 5). Away from the straits over the plateau, the currents sculpture the seafloor sediments, forming a variety of transverse and longitudinal bedforms such as obstacle marks, sand ribbons, burchan shaped dunes, sand waves, megaripples and drift deposits behind obstacles (Fig.5) (Lykousis 2001, figs 2, 7 and 10; Ferentinos & Papatheodorou 1993; Ferentinos et al. 2010). This current transports the sediments to the shelf edge and cascades them to the deep basin (Lykousis 2001; Canals et al. 2010).

**Sedimentation rates**

The average sediment thickness deposited on the erosion terrace cut across all over Hellenic shelf by the rising sea during the Last Glacial Transgression about 18 ka BP, ranges between 10 and 30 m, giving an average sedimentation rate on the shelf of between 55 and 165 cm/ka respectively. These sedimentation rates are much higher than the sedimentation rates observed in the Hellenic shelf basins, where sedimentation rates for the Holocene and for the last 18ka is between 4 and 22 cm/ka (Piper & Perissoratis 2003, table 1; Geraga et al. 2000, 2005, 2008, Giresse et al. 2003; Rousakis et al. 2004). Rousakis et al. (2004) in the North Aegean Trough observed a decrease in the sedimentation rates from the 30 cm/ka in the onset of the Last Glacial Transgression to 15.4 cm/ka during the period of the sapropel formation and to 14 cm/ka in the post-sapropel period.
The high sedimentation rates prevailing on the erosion terraces during the post Last Glacial Transgression in the shelf compared to the low sedimentation rates on the basin floors, suggest that most of the terrigenous land derived material remains on the shelf. The high sedimentation rates of about 30 cm/ka observed at the onset of the Last Glacial Transgression can be attributed to the direct discharge of river sediment to the basins rims during the Last Glacial Lowstand. Furthermore, the relatively increased rates of deposition of 15 to 25 cm/ka observed in some basins are attributed to additional sediment input by gravitational processes. The sedimentation rates in the land-locked and tectonically active Corinth and Patras Gulfs are in general higher than in the Ionian and Aegean shelves. The sediment thickness deposited on the erosion terrace is about 50 m in the last 18 ka and about 20 m in the last 12 ka in the Patras and Corinth Gulfs, giving average sedimentation rates of 2.7m/ka and 1.7m/ka respectively (Ferentinos et al. 1985; Chronis et al. 1991).

However, in the Corinth Gulf basin the sedimentation rate is between 5 and 20 times higher than in the basins of the Aegean shelf. Sedimentation rates in the Corinth basin in the last 12 ka is about 100 cm/ka whilst for the period between 12 and 250 ka BP is about 50 cm/ka (Bell et al. 2009).

Geo-hazards

Gravitative mass movements

Gravitative mass movements have been observed all over the shelf in the Aegean and Ionian seas, (Ferentinos, 1992; Lykousis & Chronis 1989; Lykousis 1991 b; Papatheodorou & Ferentinos 1993, 1997; Hasiotis et al. 2002; Lykousis et al. 2009) affecting the late Pleistocene and Holocene deposited sediments.

Three main types of slope failures were identified in the acoustically stratified late Quaternary layers: (i) sliding of masses with no or only slight internal deformation on a basal planar or concave shear surface, (ii) sliding of masses with disintegration of sediment fabric into flows and (iii) slow downslope creep.

Many of the disintegrating slides as they move downslope display a continuum of transformation from slides to debris flows, to turbidity currents.
Slides are associated with: (i) active delta fronts and delta slopes with gradients ranging from 2° to 5° (Fig. 6), (ii) active delta fronts and delta slopes with gradients ranging from 6° to 20°, (iii) inactive delta fronts with slope gradients from 2° to 5°, (iv) well stratified gentle slopes with gradients from 1° to 5° and (v) fault escarpments with slope gradients from 10° to 40° (Ferentinos & Papatheodorou 1993, fig 9).

The slides, whose thicknesses are between 5 and 25m, take place over slip planes which are: (i) bedding planes parallel to the slope with a dip from 1° to 5°, (ii) unconformities parallel to the slope with a dip from 2° to 7°, (iii) fault planes with a dip between 10° and 40° and (iv) deltaic foresets or bottom sets with a dip between 1° and 10°.

Most of the dated mass movements occurred during and after the Last Glacial Trangression (Lykousis & Chronis 1989). Some landslides occurred 2 to 3 ka BP (Stefatos et al. 2006) and some others during the last 300 years (Galanopoulos et al. 1966; Papatheodorou & Ferentinos 1997; Hasiotis et al. 2002; Lykousis et al. 2005, 2007). The latter time group of landslides occurred in the Corinth Gulf and the slides were caused by known earthquake events (Papatheodorou & Ferentinos 1997; Hasiotis et al. 2002; Lykousis 2005, 2007).

The causative mechanisms responsible for the triggering of the landslides in the Hellenic shelf, are mainly cyclic loading stresses, induced by earthquakes and, an increase in the excess pore pressure which induces sediment liquefaction. Other factors that can also contribute to sliding initiation are slope steepening through tectonic control and gas (methane) in the sediment pores (Ferentinos 1992; Lykousis 1991b; Papatheodorou & Ferentinos 1993).

Based on the existing data, the areas where most of the mass movements are observed are the Corinth Gulf, the shelf adjacent to the Lefkada–Kefallinia transform fault and the shelves bordering the North Aegean Trough. These areas coincide with the areas of maximum expected peak ground accelerations in Greece.

Active Gas Seeping
Gas in the sediment pores has been observed all over the Hellenic shelf (Papatheodorou et al. 1993; Soter 1999; Hasiotis et al. 1996; Chistodoulou et al. 2003).

The gas charged sediments are found in Pleistocene and present-day land-locked gulfs, Pleistocene and present day deltaic environments, open sea shelf environments and lakes. The gas is usually associated with the: i) Holocene/Pleistocene(?) boundary which seems to be an accumulative horizon (Fig. 7); (ii) dipping Pleistocene or Pliocene bedding planes through which the gas migrates up-dip, and (iii) fault planes through which the gas migrates (Fig. 7). The gas is also associated with morphological features either on the seabed, or under the seabed, such as surface and buried pockmarks (Fig. 8), intra sedimentary and seabed dome shaped mounds, gas-pockets and cemented sediments on the seafloor.

The gas found in the Pleistocene and present-day land-locked gulfs and deltaic environments is of biogenic origin, while that found in the pre-Quaternary open sea shelf environments which are associated with faults and salt doming, is probably of thermogenic origin.

Detailed studies in two pockmark fields in Patras and Corinth gulfs respectively, and in a gas seeping field over a salt dome in Katakolo, (Ionian Sea shelf) have shown that (i) the Corinth Gulf pockmark field was formed by ground water seepage whilst the Patras Gulf one, by biogenic methane seepage (Christodoulou et al. 2003; Christodoulou 2010), (ii) the methane seeping from the seafloor in the Katakolo gas seeping field is of thermogenic origin (Etioppe et al. 2006; Christodoulou 2010), (iii) the methane concentration in the water near the seabed inside the pockmarks ranges between 0.4 and 1.1 μmol/l over a background concentration between 0.002 and 0.2 μmol/l (Christodoulou et al. 2003; Christodoulou 2010), (iv) the estimated contribution of methane to the atmosphere from the pockmark field is between 0.09 and 0.35 t/year (v) the methane concentration in the water near the seabed in the Katakolo field is over 4.5 μmol/l whilst the estimated contribution of methane to the atmosphere is between 1262 and 1500 t/year (Christodoulou 2010). The pockmark field in Patras gulf, which was activated twice by earthquakes in 1993 and 2008 (Fig. 8), had an impact in the stability of the breakwater and imposed in general constraints in the construction of
Patras harbour (Chistodoulou 2010). Similarly the gas seepage in Katakolo is considered hazardous to human activities and health due to the presence of methane and hydrosulfide in levels that are potentially inflammable and toxic (Etioppe et al. 2006).

**Tsunami**

The study of Tsunami Catalogues reveals a high frequency of tsunami occurrence over the Hellenic shelf. A total of 160 probable and definite tsunami events have recorded since 1628 AD. The two most recent destructive tsunami in the Hellenic shelf occurred in 1956 in Amorgos Island and 1963 in western Corinth Gulf (Papadopoulos & Chalkis 1984). The causal mechanisms of the former tsunami were co-seismic displacement of the seafloor and triggering of submarine landslide (Perissoratis & Papadopoulos 1992) whilst the causal mechanism of the latter tsunami was a coastal submarine landslide (Stefatos et al. 2006). A tsunami risk assessment study in the Corinth Gulf, a tsunamigenic zone (Papadopoulos & Chalkis 1984) based on a wealth of offshore geological data have shown that: (i) a maximum probable 6.7 (Mw) offshore earthquake would produce a rapture in the seafloor with a maximum displacement of 1.08 m and a potential wave of the same height (ii) a submarine mass failure would have the potential of generating tsunami with maximum wave height of 4.04 m over the source (Stefatos et al. 2006).

**Conclusions**

The Hellenic shelf is located within one of the world’s most seismically active areas and has experienced extreme tectonism throughout Quaternary time. These tectonic forces are responsible for: (i) the overall configuration of the Hellenic shelf as well as for the formation and orientation of the basins within it, (ii) the spatial and temporal variability in the uplift and subsidence rates observed in the Hellenic shelf, (iii) the sediment yields and sedimentary rates and (iv) the frequent occurrence of a variety of offshore geological hazards. Sea-level changes that occurred throughout the last 400ka has also played an important factor in the configuration of the Hellenic shelf with the formation of
(i) transgressive and regressive depositional sequences and (ii) the transport and deposition of sediment over the shelf.

References


Figure captions

Fig. 1 Regional map of the Hellenic Arc shelf showing the bathymetric and active tectonic features (compiled from: Bartole et al. (1983); Brooks and Ferentinos (1984); Mascle & Martin (1990); Ferentinos 1992; Lykousis et al. (1995); Koukouvelas & Aydin (2002); Piper & Perissoratis (2003), Kokkalas et al. (2006); Kokkalas & Aydin 2012). HSz: Hellenic Subduction zone; LKe: Lefkada-Kefalonia escarpment; IFb: Ilyrian folding belt; NAT: North Aegean Trough; CSb: Cretan Sea fore-arc basin; ACp: Attico-Cyclades platform; S-Lp: Sporades-Limnos platform; S-Et: Skyros-Edremit trough; Sab: Saros basin; Spb: Sporades basin; Thr: Thermaikos Gulf, M-Ts: Macedonian-Thracian shelf, MG: Messinianos gulf; LG: Lakonikos gulf; AG: Argolikos gulf; PG: Patras Gulf; CG: Corinth Gulf; Ikb: Ikaria Basin; Amb: Amorgos Basin; Anb: Anafi Basin and Kob: Kos Basin.


1b. Map showing the bathymetry and fault lines in the Sporades basin, based on two marine geophysical surveys carried out in 1978 (RSS Shackleton cruise) and 1982 (Discovery cruise) by the Department of Oceanography, University of Swansea, Wales. S-Afz: Sithonia/Athos fault zone, S-Lfz: Sporades/Limnos fault zone.

1c. Map showing the bathymetry and fault lines in the Maleas-Kythera-Crete Ridge, based on a marine survey carried out in 1988 by the Laboratory of Marine Geology and Physical Oceanography, University of Patras, Greece. Grb: Gramvousa basin, Kb: Kissamos basin.

Fig. 2 Sparker profile across the Gramvousa and Kissamos basins in western Crete (for the profile location see Fig. 1c). The overall configuration pattern the faults exhibit on the sparker profile indicates oblique normal faulting. Profile was collected during the RRS Shackleton 1972 and 1974 cruises in the Cretan Sea.

Fig. 3 High resolution 3.5kHz seismic profile across the Sithonia/Athos fault zone (for profile location see Fig.1b) displaying Holocene/Upper Quaternary layered sediments, which are affected by thrusts faulting, crumbled against the Sithonia/Athos fault zone indicative of active transpressional forces.

Fig. 4 3.5 kHz O.R.E seismic profile across an anticlinical structure formed between two antithetic listric faults (for the profile location see Fig. 1b), showing the formation of a graben at the crest of the anticline and upslope facing antithetic growth faults in the flank. The approximate 30m thick sedimentary sequence that drapes the slope corresponds to hemipelagic sedimentation. The upper three meters of the sequence is deposited during the late Holocene Highstand whilst the rest is deposited during the 2.2 IOS Lowstand.
Fig. 5 Multibeam seafloor images (for location see Fig. 1) showing the formation of a sand drift deposit at the lee side of a ridge due to the overflow of bottom currents over the ridge crest. The lineations on the surface of the sand deposits correspond to erosional furrows behind obstacles indicating bottom flow direction. RC: Ridge crest; SD: Sand drift deposits; F: Furrows.

Fig. 6 Side scan sonar mosaic (a) and 3.5KHz seismic profiles in the dip direction (C1) and in the strike direction (C2) along the Selinountas and Keranitis rivers delta-front in the Corinth Gulf (for location see Fig. 1) showing the areal extend of a complex translational slide less than 5m thick with lateral deformational complexity. The slide covers a seafloor area of about 600,000 m² and occurred in 1995 by a 6.2-R earthquake Ez: Evacuation zone; Spl: Slide plane; Mdl: Main depositional lobe; Hdz: High deformational zone; Ldz: Low deformational zone; Ndz: Non-deformational zone; PR: Pressure ridges; Dd: distal deposits; ATZ: Acoustic Turbidity Zone indicating the presence of bubble phase gas and GPL: gas plumes.

Fig. 7 3.5KHz (a) and air-gun (b) seismic profiles across active faults in the Patras Gulf (for location of the profiles see inset showing: (i) gas accumulation in the Pleistocene (?)–Holocene interface (ATZ: Acoustic Turbid Zone; (ii) active faults (F) displacing the Pleistocene (?)–Holocene interface and the seafloor and (iii) gas plumes (GP/GPl) rising along the fault planes to the Holocene cover forming domes (D), and to the water column.

Fig. 8 Detailed bathymetric map (a) showing a large pockmark field off Patras Harbour in the Patras Gulf (for location see inset. The side scan sonar record (c) shows gas plumes (GPl) rising from the pockmarks to the water column about 24 hours after a nearby 6.2R earthquake in 2008.
Figure 2

Gramvousa Ridge

Gramvousa Basin

WSW

Kissamos Basin

ENE

Spatha Ridge

0.1 sec TWTT

0 1 2 km